SEASONAL AND INTERANNUAL VARIABILITY IN ATMOSPHERIC TURBIDITY OVER SOUTH AFRICA

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ABSTRACT

Aerosols affect climate by attenuating solar radiation and acting as cloud condensation nuclei. Despite their importance in the climate system, our understanding of the time-space variability of aerosols is fragmentary. Measurements and reliable estimates of atmospheric turbidity—the total column amount of aerosol—are scarce in most countries and this is especially true in the Southern Hemisphere. Very little is known about the seasonal, interannual and spatial variability of aerosols over the southern half of the globe.

In this paper, we estimate monthly averaged atmospheric turbidity from surface climate data at eight locations in South Africa, regardless of cloud cover. Findings include new estimates of turbidity trends and variability over South Africa. Seasonal trends are evident at many stations, although there is no consistent trend. Over recent decades, turbidity has generally been stable at six of the eight stations. Our methodology can be applied at any location where the requisite climate data are available and, therefore, holds promise for a more complete, and possibly global, climatology of aerosols. Copyright © 2001 Royal Meteorological Society.

KEY WORDS: aerosols; turbidity; climate; radiation; modelling; South Africa

1. INTRODUCTION

Quantifying the processes by which aerosols affect climate is crucial to understanding how humans influence climate and how we may alter climate in the future; this is especially important as anthropogenic emissions of aerosols continue to increase (Penner *et al.*, 1994; Wolf and Hidy, 1997; Kiehl, 1999). Aerosols are known to affect climate in two ways. They influence the planetary albedo by scattering and absorbing radiation (direct effect) and they modify the physical and radiative properties of clouds by acting as cloud condensation nuclei (indirect effect). Although the exact magnitude of aerosol-induced radiative forcing is uncertain, aerosols are thought to have a net cooling effect. Thus, they may mask the warming effects of anthropogenic increases in greenhouse gases (Wigley, 1989; Kaufman *et al.*, 1991; Charlson *et al.*, 1992; Penner *et al.*, 1992; Christopher *et al.*, 1996; Kiehl, 1999). Indeed, aerosols may have delayed the detection of global warming due to greenhouse gases (Wigley and Raper, 1992).

A key reason for the uncertainty about aerosol–climate interactions is that aerosols demonstrate high variability, both spatially and temporally, and the extent of their variability has not been fully investigated. There is a paucity of reliable data on the amount of aerosol in the atmosphere. Conventional techniques for quantifying atmospheric turbidity—the amount of aerosol integrated vertically through the atmosphere—typically require costly and sophisticated instrumentation and clear-sky conditions (e.g. Dutton *et al.*, 1994; Holben *et al.*, 1998). Recent satellite estimates of turbidity are also restricted to clear-sky conditions and non-terrestrial regions (Tanré *et al.*, 1996; Husar *et al.*, 1997). The clear-sky restriction produces gaps in the turbidity record, as well as a possible bias because cloudless skies are associated with a limited set of synoptic situations.

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In addition to these potential biases, atmospheric turbidity measurements and estimates are not optimally distributed in space. In other words, turbidity studies are geographically biased. With only a few, mostly recent, exceptions (Shaw, 1982; Herber *et al.*, 1991; Villevalde *et al.*, 1994; Herber *et al.*, 1996; Yatko and Garrison, 1996; Holben *et al.*, 1998; Veefkind and de Leeuw, 1998), most studies have focused on Northern Hemisphere turbidities (e.g. Sadler, 1978; Cerf, 1980; Freund, 1983; Prodi *et al.*, 1984; Al-Jamal *et al.*, 1987; Gueymard, 1994; Holben *et al.*, 1996; Jacovides *et al.*, 1996a,b; Adeyefa *et al.*, 1997; Blumthaler *et al.*, 1997; Maduekwe and Chendo, 1997; Pinker *et al.*, 1997; Gueymard and Garrison, 1998; Kambezidis *et al.*, 1998; Chiapello *et al.*, 1999; Russell *et al.*, 1999; Veefkind *et al.*, 1999; von Hoyningen-Huene *et al.*, 1999). Given that there are distinct differences in aerosol sources, emission rates and sampling between the Southern and Northern Hemispheres—the majority of anthropogenic aerosols are emitted in the Northern Hemisphere (Penner *et al.*, 1994; Trenberth, 1996; Wolf and Hidy, 1997)—there has been a Northern Hemisphere sampling bias which has led to questionable inferences about atmospheric turbidity over large parts of the Southern Hemisphere.

Owing to such temporal and spatial constraints, the turbidity record contains data for too few locations, for mostly short time periods and primarily for the Northern Hemisphere. There appears to be no reliable, long-term record of atmospheric turbidity for determining trends in aerosol variability, especially in the Southern Hemisphere.

The present investigation employed a recently developed model to estimate monthly averaged turbidity from surface climate data at eight locations in South Africa, regardless of cloud cover. The temporal extent of the South African historical climate record provided the opportunity to examine both seasonal and long-term variability in turbidity in the Southern Hemisphere. This methodology makes use of available climate data—and is not restricted to clear-sky conditions—which allows the estimation and presentation of a previously unavailable climatology of turbidity.

2. ATMOSPHERIC TURBIDITY MODEL

In addition to aerosol attenuation, solar radiation passing through the atmosphere is attenuated by molecular scattering and absorption by ozone, nitrogen dioxide, water vapour, carbon dioxide and oxygen. In the absence of clouds, surface irradiance can be modelled as a function of these attenuation processes. Many existing models then can be inverted to obtain the aerosol contribution wherever radiation is measured and the other components are measured or adequately estimated. Most of these models, however, can only estimate *instantaneous* turbidity since they are dependent on optical air mass (a function of the position of the sun) and cannot be analytically integrated over time.

Power (1999, 2001a) significantly reduced this limitation by using a physically based, high-resolution spectral radiation model (Gueymard, 1995) to parameterize clear-sky beam irradiance with polynomial expansions of the sine of the solar-elevation angle. This structure permits analytic integration and inversion and, in turn, provides an expression for estimating daily averages of Ångström's turbidity coefficient¹ (β). The model is

$$\hat{\beta} = c_1^{-1} \ln \left(\frac{H_{bc}}{k E_{0n} c_0 \exp(c_2 \bar{w}) \bar{u}_o^{c_3} \left(\frac{\bar{p}}{p_0} \right)^{c_4} \bar{u}_n^{c_5} \sum_{i=0}^7 \varphi_i J_{i+1}} \right), \quad (1)$$

where c_i and φ_i are coefficients obtained from a linear least-squares fit to the reference data (Appendix A), H_{bc} is the 'potential' daily clear-sky beam irradiation on a horizontal surface (J m^{-2} ; see below), $k = 24/\pi$ (h), E_{0n} is the broadband extraterrestrial irradiance at normal incidence and actual earth–sun distance (W m^{-2}), w is precipitable water (cm), u_o is total column ozone (atm-cm), p is atmospheric pressure (mb), p_0 is the sea-level standard pressure (mb), u_n is total column nitrogen dioxide (atm-cm) and the overbars denote the daily averages of each of the respective quantities. The term J_i represents

$$J_i = \int_0^{\omega_0} \sin^i h \, d\omega, \quad (2)$$

where h is the solar-elevation angle, ω is the hour angle and ω_0 is the hour angle at sunrise. In Equation (1), a solar-angle correction factor ($\sin h$) is incorporated in J_{i+1} . Analytic solutions to Equation (2) were initially proposed by Gueymard (1993) and appear in Appendix B.

Potential daily clear-sky beam irradiation on a horizontal surface (H_{bc}) is the beam radiation that would have been incident *if* the sky had been clear all day. Thus, it is a theoretical, ‘potential’ quantity that is only ever observed on days that are cloud free. Following Louche *et al.* (1991), Gueymard (1993) and Power (2001b), monthly averages of the potential daily clear-sky beam irradiation ($\bar{H}_{bc}$) can be estimated from

$$\bar{H}_{bc} = \bar{H}_b \bar{S}(\bar{s})^{-1} \quad (\text{J m}^{-2}), \quad (3)$$

where H_b is the *observed* daily beam irradiation (J m^{-2}), S is the astronomical day duration (daylength) and s is the duration of bright sunshine.² In Equation (3), the overbars denote the monthly average values of each of the respective quantities. Because Equation (3) provides monthly estimates of potential irradiation, forcing Equation (1) with monthly averaged climate data permits monthly averaged turbidity estimates. In this way, turbidity can be estimated regardless of cloud cover.

Power (1999, 2001a) evaluated Equation (1) (using Equation (3) to estimate $\bar{H}_{bc}$) with monthly turbidity and climate data from five stations in Europe and North America. Willmott’s absolute-difference-based index of agreement (d) (Willmott *et al.*, 1985)³ was 0.70, while the mean-absolute error was 0.02 for a mean monthly averaged β of 0.09.

3. ESTIMATES OF SOUTH AFRICAN TURBIDITY

Estimating monthly averages of Ångström’s turbidity coefficient ($\bar{\beta}_m$) from Equations (1) and (3) requires monthly averages of H_b , s , w , u_o , p and u_n . Values of E_{0m} , ω_0 and S can be easily calculated, for any location, from established astronomical relationships.

3.1. Climate forcing

In the Southern Hemisphere, there are few locations where the requisite climate data are available. The main constraint is the availability of accurate solar radiation observations. South Africa, however, offers a well-maintained network of first-order weather stations. The South African Weather Bureau (SAWB) provided requisite data for eight stations in South Africa. Four of the stations (Alexander Bay, Cape Town, Durban and Port Elizabeth) are coastal, while Bloemfontein, Grootfontein, Pretoria and Upington are located inland at elevations above 800 m (Table I, Figure 1). A temporal ‘window’ common to the

Table I. Latitude, longitude, elevation, period and length of record, and population for each of the eight climate stations for which monthly estimates of Ångström’s turbidity coefficient (β) were made

	Latitude (S)	Longitude (E)	Elevation (m)	Period	Length (years)	Population
Alexander Bay	28.57	16.53	21	1959–1984	26	< 100 000
Bloemfontein	29.10	26.30	1351	1959–1993	35	300 000
Cape Town	33.97	18.60	44	1966–1995	30	2 350 000
Durban	29.96	30.95	8	1978–1991	14	1 130 000
Grootfontein	31.48	25.03	1270	1969–1990	22	< 100 000
Port Elizabeth	33.98	25.60	60	1959–1991	33	853 000
Pretoria	25.73	28.18	1330	1965–1997	33	1 080 000
Upington	28.40	21.27	836	1969–1992	24	< 100 000

Population figures are 1991 estimates (United Nations, 1999) for the greater metropolitan area of each location.

Figure 1. Relative elevation in Southern Africa (grey-scale shading), with the locations of the SAWB climate stations for which the requisite data for making turbidity estimates were available

station data prescribed the time period over which monthly estimates of turbidity could be made for each station. Overall record lengths average 27 years and range between 14 and 35 years (Table I).

As input to Equation (3), $\bar{H}_b$ was computed as the difference between the monthly averaged global and diffuse components. Both the global and the diffuse irradiances were measured with Kipp and Zonen CM5 pyranometers, and the SAWB applied shadowband correction factors to the diffuse data. The uncertainty associated with the monthly averaged global and diffuse irradiation is estimated at less than 2% (D. Esterhuyse, personal communication, 2000). Bright sunshine duration (s) was measured with Campbell–Stokes sunshine recorders. These instruments have an expected root-mean-square error of approximately 4.1% of the monthly percentage of possible sunshine (Benson *et al.*, 1984).

As input to Equation (1), mean monthly estimates of w were computed using Gueymard's (1994) algorithm as a function of monthly averaged air temperature and monthly averaged relative humidity. Monthly averaged u_o was estimated, using Van Heuklon's (1979) algorithm, from latitude, longitude and the climatologically average day of the month.⁴ Monthly averaged u_n was estimated according to the size (population) of each location.⁵ A value of 2.0 matm-cm was assigned to those cities with a population over 1 million, 1.0 matm-cm was used for cities with populations between 100 000 and 1 million, and a value of 0.15 matm-cm was assigned for the remaining locations.

3.2. Seasonal variability

The most striking result is the relatively low levels of turbidity in South Africa (Table II); long-term averages of $\hat{\beta}_m$ range between 0.013 and 0.047. Highest turbidity occurs in the main centres (Pretoria, Durban and Cape Town), while the lowest turbidity occurs in the small township of Alexander Bay located on the Atlantic coast near the Namibian border. In the larger cities, anthropogenic aerosols may be augmenting the turbidity signal.

Many stations exhibit marked seasonal trends in turbidity. Bloemfontein, Grootfontein and Pretoria (Figure 2), for example, exhibit typical summer highs and winter lows in turbidity that have been observed elsewhere (e.g. Peterson and Flowers, 1977; Uboegbulam and Davies, 1983; Louche *et al.*, 1987; Garrison and Sahami, 1995). These three stations are located in the mountains at elevations of 1351, 1270 and 1330 m, respectively (Table I, Figure 1). In the summer, they are probably affected by low-pressure systems that draw in moist air from the Indian Ocean. Higher precipitable water in the summer (Figure 3) can enhance hygroscopic aerosol growth and may contribute to the summer highs in turbidity at these three locations.

Table II. Long-term monthly and annual averages and standard deviations (in parentheses) of Ångström's turbidity coefficients estimated at each of the eight climate stations in South Africa

	AB	BL	CT	DN	GR	PE	PR	UP
January	0.007 (0.003)	0.026 (0.014)	0.021 (0.010)	0.021 (0.014)	0.022 (0.013)	0.021 (0.018)	0.059 (0.021)	0.02 (0.011)
February	0.007 (0.004)	0.025 (0.013)	0.021 (0.011)	0.017 (0.009)	0.027 (0.014)	0.021 (0.012)	0.071 (0.023)	0.028 (0.013)
March	0.004 (0.003)	0.018 (0.011)	0.016 (0.011)	0.025 (0.019)	0.013 (0.008)	0.021 (0.017)	0.058 (0.020)	0.015 (0.009)
April	0.006 (0.006)	0.015 (0.009)	0.020 (0.018)	0.019 (0.007)	0.011 (0.007)	0.019 (0.015)	0.044 (0.020)	0.018 (0.011)
May	0.010 (0.009)	0.010 (0.005)	0.032 (0.024)	0.030 (0.011)	0.019 (0.017)	0.019 (0.014)	0.031 (0.013)	0.015 (0.015)
June	0.017 (0.019)	0.013 (0.014)	0.030 (0.023)	0.038 (0.027)	— —	0.018 (0.020)	0.027 (0.013)	0.016 (0.012)
July	0.013 (0.011)	0.008 (0.006)	0.030 (0.021)	0.036 (0.012)	0.005 (0.004)	0.020 (0.020)	0.031 (0.012)	0.013 (0.009)
August	0.017 (0.013)	0.016 (0.013)	0.039 (0.019)	0.043 (0.013)	0.013 (0.008)	0.030 (0.022)	0.039 (0.012)	0.025 (0.010)
September	0.018 (0.013)	0.028 (0.017)	0.038 (0.020)	0.033 (0.011)	0.023 (0.016)	0.036 (0.025)	0.047 (0.015)	0.027 (0.013)
October	0.014 (0.010)	0.031 (0.016)	0.034 (0.013)	0.028 (0.012)	0.026 (0.014)	0.033 (0.018)	0.053 (0.016)	0.032 (0.008)
November	0.010 (0.006)	0.028 (0.013)	0.032 (0.015)	0.026 (0.014)	0.023 (0.013)	0.035 (0.015)	0.055 (0.016)	0.027 (0.012)
December	0.008 (0.006)	0.022 (0.020)	0.024 (0.012)	0.025 (0.025)	0.021 (0.012)	0.023 (0.014)	0.052 (0.017)	0.021 (0.015)
Annual	0.013 (0.011)	0.022 (0.015)	0.028 (0.018)	0.029 (0.016)	0.020 (0.014)	0.025 (0.019)	0.047 (0.021)	0.022 (0.013)

AB = Alexander Bay; BL = Bloemfontein; CT = Cape Town; DN = Durban; GR = Grootfontein; PE = Port Elizabeth; PR = Pretoria; UP = Upington.

Figure 2. Boxplots of monthly estimates of Ångström's turbidity coefficient ($\hat{\beta}_m$) in Pretoria 1965–1997. Bloemfontein and Grootfontein also exhibited summer maxima and winter minima. Asterisks represent outliers in the data

Figure 3. Long-term averages of monthly estimates of precipitable water (w) at each of the eight climate stations. Stations at higher elevations (Uppington, Grootfontein, Pretoria and Bloemfontein) are relatively dry, while the maritime environments of Durban, Port Elizabeth, Cape Town and Alexander Bay are characterized by higher atmospheric moisture

However, precipitable water is only one of many factors that can influence turbidity at any given location. Other climatic factors include convective activity, seasonal differences in air mass origin and the synoptic situation, as well as dispersion, transformation and removal processes. Indeed, precipitable water is highest in summer at *all* locations (Figure 3), yet Durban (Figure 4) and Alexander Bay, for example, exhibit peak turbidity in the winter.

Superimposed on the climatic influences is variability in the emissions of aerosols, particularly anthropogenic aerosols, and this further complicates interpretation of the aerosol signal. At stations with a summer maximum in turbidity (Bloemfontein, Grootfontein and Pretoria), it may be the ambient aerosol load in the atmospheric column that is being detected. At stations with other seasonal maxima, anthropogenic aerosols in the boundary layer—which are typically highest in winter—may dominate the aerosol signal. The climate station at Bloemfontein, for example, where turbidity is highest in summer, is located approximately 23 km east of the city where anthropogenic aerosols may play a lesser role in determining turbidity. Durban, in contrast, suffers from air pollution problems in the winter (D. Esterhuyse, personal communication, 2000) and this may contribute to the observed winter maxima in turbidity (Figure 4).

Turbidity in Cape Town is at a maximum in late winter and spring (Figure 5). With a population of over 2 million people, anthropogenic aerosols from the city may well influence turbidity at this station. Depending on the synoptic situation, maritime aerosols may also influence the turbidity signal in Cape

Figure 4. Boxplots of monthly estimates of Ångström's turbidity coefficient ($\hat{\beta}_m$) in Durban 1978–1991. Alexander Bay also demonstrated winter maxima. Asterisks represent outliers in the data

Figure 5. Boxplots of monthly estimates of Ångström's turbidity coefficient ($\hat{\beta}_m$) in Cape Town 1966–1995. Asterisks represent outliers in the data

Town, as well as in the other coastal stations in Alexander Bay, Port Elizabeth and Durban. Air masses originating from the Kalahari Desert may contribute to the aerosol load in Pretoria, Bloemfontein, Upington and Grootfontein. Clearly, there are many factors—both climatic and non-climatic—that determine turbidity patterns at any given location. It is difficult to attribute seasonal trends in turbidity to any one causal factor.

In an effort to quantify seasonal trends common among the eight stations, Pearson correlation coefficients (r) were computed between the long-term monthly medians of the turbidity estimates at all stations⁶ (Table III). This analysis corroborated the shared seasonal trends described above. Summer maxima in Bloemfontein and Pretoria, for instance, are reflected in the high r value for these two stations (0.78). Of all the interstation comparisons, the highest positive correlations (for instance, those where $r \geq 0.7$) occur between coastal stations *or* between upland stations, with the exception of Port Elizabeth (coastal) and Bloemfontein (upland), and Upington (upland) and Port Elizabeth. Furthermore, a moderately high negative correlation ($r = -0.62$) exists between Pretoria, in the mountains and Durban, on the east coast. These zonal associations may reflect the influence of synoptic climatology (e.g. dominant air masses or trajectories) on the aerosol climatology.

3.3. Interannual variability and trends

In order to examine long-term trends in turbidity over South Africa, moving averages of $\hat{\beta}_m$ were computed at each station. Although interannual variability in turbidity is evident at all stations (Table II), at most stations turbidity is generally stable through the respective time series. However, a linear model

Table III. Pearson correlation coefficients (r) between the long-term monthly medians of the ($\hat{\beta}_m$) estimates at the eight climate stations in South Africa

	AB	BL	CT	DN	GR	PE	PR
BL	−0.010						
CT	0.858	0.225					
DN	0.713	−0.323	0.740				
GR	0.119	0.790	0.225	−0.336			
PE	0.576	0.714	0.804	0.360	0.538		
PR	−0.437	0.782	−0.227	−0.617	0.574	0.285	
UP	0.503	0.699	0.554	0.061	0.584	0.755	0.537

Key as for Table II.

fit to the Cape Town and Grootfontein time series both show long-term trends in turbidity. In Cape Town, where the long-term averaged $\hat{\beta}_m$ is 0.028, $\hat{\beta}_m$ has, on average, increased by approximately 0.022 between 1966 and 1995 (Figure 6). A simultaneous 0.47 cm average increase in monthly precipitable water (w) (Figure 7) may be a contributing factor to the observed trend. Meanwhile, turbidity in Grootfontein appears to have decreased on average by 0.013 during the period 1969–1990 (Figure 8). Interestingly, a concomitant 0.80 cm average *increase* in monthly w (Figure 9) suggests that the observed decrease in turbidity is probably the result of changes in other climatic factors and/or a reduction in aerosol emissions.

In Cape Town, a period of steadily increasing turbidity commenced in mid-1991 and continued until the end of 1994 (Figure 6). Similar temporal trends are evident in Pretoria (Figure 10) and Upington (not shown). Aerosols ejected from the June 1991 eruption of Mount Pinatubo may have contributed to these short-term trends. There is, however, no evidence of a post-Pinatubo increase in aerosols in Bloemfontein, while in Durban, Port Elizabeth, Grootfontein and Alexander Bay there are insufficient turbidity data after the eruption to enable the assessment of any possible effects on the aerosol climatology. A noticeable short-term increase in turbidity, commencing in 1963, is evident in Port Elizabeth (Figure 11), Alexander Bay and Bloemfontein (not shown). These trends may, in part, be attributable to the March 1963 eruption of Mount Agung in Indonesia. Such trends cannot be observed at other stations because the time series

Figure 6. Linear trend model fit to monthly estimates of Ångström's turbidity coefficient ($\hat{\beta}_m$) for Cape Town 1966–1995. The model is $\hat{\beta}_i = (1.77 \times 10^{-2}) + (6.00 \times 10^{-5} \times t_i)$ where $\hat{\beta}_i$ is the estimated value of Ångström's turbidity coefficient for the i th month and t_i is the i th month since January 1966

Figure 7. Linear trend model fit to monthly estimates of precipitable water (w) for Cape Town 1966–1995. The model is $w_i = 2.24 + (1.3 \times 10^{-3} \times t_i)$, where w_i is the estimated value of precipitable water for the i th month and t_i is the i th month since January 1966

Figure 8. Linear trend model fit to monthly estimates of Ångström's turbidity coefficient ($\hat{\beta}_m$) for Grootfontein 1969–1990. The model is $\hat{\beta}_i = (2.50 \times 10^{-2}) - (5.06 \times 10^{-5} \times t_i)$, where $\hat{\beta}_i$ is the estimated value of Ångström's turbidity coefficient for the i th month and t_i is the i th month since January 1969

Figure 9. Linear trend model fit to monthly estimates of precipitable water (w) for Grootfontein 1969–1990. The model is $w_i = 1.44 + (3.02 \times 10^{-3} \times t_i)$, where w_i is the estimated value of precipitable water for the i th month and t_i is the i th month since January 1969

Figure 10. Monthly estimates and 12-monthly moving averages (solid line) of Ångström's turbidity coefficient ($\hat{\beta}_m$) in Pretoria 1965–1997. The rise in turbidity commencing in 1991 was also evident in Cape Town and Upington

Figure 11. Monthly estimates and 12-monthly moving averages (solid line) of Ångström's turbidity coefficient ($\hat{\beta}_m$) in Port Elizabeth 1959–1991. The peak in turbidity commencing in 1963 was also evident in Alexander Bay and Bloemfontein

begin later than 1963 (Table I). The 1982 eruption of El Chichón does not appear to be detectable in any of the turbidity signatures.

It appears that the majority of stations show no evidence of any long-term changes in turbidity. Furthermore, because turbidity patterns are not consistent across all stations, the long-term trends that do exist (in Cape Town and Grootfontein) are probably the result of local or regional changes in climate or in aerosol emissions (natural or anthropogenic). The long-term trend identified in Cape Town should also be interpreted with caution; without the 1991–1994 increase, which may be due to the Pinatubo eruption, turbidity would appear to be relatively stable. As with the seasonal trends identified above, it is difficult to determine the relative importance of climatic changes versus emission patterns on long-term trends and variability in turbidity.

4. CONCLUSIONS

This research provides a preliminary climatology of aerosols previously unavailable for South Africa. Compared with the Northern Hemisphere, turbidity in South Africa was found to be very low. The highest turbidities are in the larger cities of Cape Town, Durban and Pretoria, while the smaller townships are characterized by relatively low turbidity. Seasonal trends are evident at many stations, although there is no consistent trend. Over recent decades, turbidity has generally been stable at most stations in South Africa. Exceptions include Cape Town, where there has been an increase in turbidity over the last 30 years, and Grootfontein, where turbidity appears to have decreased over the 25-year record. The eruptions of Mount Pinatubo and Mount Agung appear to be detectable in the turbidity signatures at some stations.

Although the present application is limited to South Africa, atmospheric turbidity can be estimated, regardless of cloud cover, wherever the requisite climate data are available. The methodology, therefore, offers the potential for more extensive aerosol climatologies than are currently available.

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APPENDIX A. COEFFICIENTS c_i AND φ_i

i	c_i	φ_i
0	0.874744	0.023009
1	-3.764677	3.718631
2	-0.053730	-6.209335
3	-0.006222	4.840892
4	-0.079219	-0.670249
5	-0.001977	-1.406449
6		0.936992
7		-0.159847

APPENDIX B. SOLUTIONS TO THE $J_i = \int_0^{\omega_0} \sin^i h \, d\omega$ INTEGRALS

$J_0 = \omega_0$

$J_1 = q(\sin \omega_0 - \omega_0 \cos \omega_0)$

$J_2 = q^2 \left[\omega_0 \left(\frac{1}{2} + \cos^2 \omega_0 \right) - \frac{3}{4} \sin 2\omega_0 \right]$

$J_3 = q^3 \left[\sin \omega_0 \left(\frac{2}{3} + \frac{11}{6} \cos^2 \omega_0 \right) - \omega_0 \cos \omega_0 \left(\frac{3}{2} + \cos^2 \omega_0 \right) \right]$

$J_4 = q^4 \left[\omega_0 \left(\frac{3}{8} + 3 \cos^2 \omega_0 + \cos^4 \omega_0 \right) - \sin 2\omega_0 \left(\frac{55}{48} + \frac{25}{24} \cos^2 \omega_0 \right) \right]$

$J_5 = q^5 \left[\sin \omega_0 \left(\frac{8}{15} + \frac{607}{120} \cos^2 \omega_0 + \frac{137}{160} \cos^4 \omega_0 \right) - \omega_0 \cos \omega_0 \left(\frac{15}{8} + 5 \cos^2 \omega_0 + \cos^4 \omega_0 \right) \right]$

$J_6 = q^6 \left[\omega_0 \left(\frac{5}{16} + \frac{45}{8} \cos^2 \omega_0 + \frac{15}{2} \cos^4 \omega_0 + \cos^6 \omega_0 \right) - \sin 2\omega_0 \left(\frac{231}{160} + \frac{91}{20} \cos^2 \omega_0 + \frac{49}{20} \cos^4 \omega_0 \right) \right]$

$J_7 = -\frac{1}{8960} [q^7 (171\,500\omega_0 \cos \omega_0 + 61\,740\omega_0 \cos 3\omega_0 + 6860\omega_0 \cos 5\omega_0 + 140\omega_0 \cos 7\omega_0$

$\quad - 42\,875 \sin \omega_0 - 48\,363 \sin 3\omega_0 - 9947 \sin 5\omega_0 - 363 \sin 7\omega_0)]$

$J_8 = \frac{1}{35\,840} [q^8 (686\,000\omega_0 + 878\,080 \omega_0 \cos 2\omega_0 + 219\,520\omega_0 \cos 4\omega_0 + 17\,920\omega_0 \cos 6\omega_0$

$\quad + 280\omega_0 \cos 8\omega_0 - 395\,136 \sin 2\omega_0 - 208\,554 \sin 4\omega_0 - 28\,544 \sin 6\omega_0 - 761 \sin 8\omega_0)],$

where $q = \cos \phi \cos \delta$, ϕ is the latitude and δ is the solar declination.

NOTES

- 1. Ångström’s turbidity coefficient (β) is a dimensionless measure of turbidity which is defined as the aerosol optical depth at a wavelength of 1.0 μm . Values of β typically range from 0.02, for a low aerosol load, to 0.5, for a high aerosol load.
- 2. Bright sunshine duration (s) is defined as the number of hours per day that the sunshine intensity exceeds some predetermined threshold of brightness.

3. Willmott's index of agreement (d) is a dimensionless statistic that varies between 0.0 and 1.0, where a value of 1.0 indicates perfect agreement between two variables and 0.0 describes complete disagreement.
4. In estimating monthly u_o , and calculating E_{om} , ω_0 and S , the climatologically average day of the month was based on the average solar declination and earth–sun distance for the month.
5. Population data were 1991 estimates (United Nations, 1999). See Table I.
6. Median values were used, rather than means, to avoid the possible influence of outliers in the data.

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